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THE HYDROGEOLOGY OF KILAUEA VOLCANO

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Abstract—The hydrogeology of Kilauea volcano and adjacent areas has been studied since the turn of this century. However, most studies to date have focused on the relatively shallow, low-salinity parts of the ground-water system, and the deeper hydrothermal system remains poorly understood. The rift zones of adjacent Mauna Loa volcano bound the regional ground-water flow system that includes Kilauea, and the area bounded by the rift zones of Kilauea and the ocean may comprise a partly isolated subsystem. Rates of ground-water recharge vary greatly over the area, and discharge is difficult to measure, because streams are ephemeral and most ground-water discharges diffusely at or below sea level. Hydrothermal systems exist at depth in Kilauea's east and southwest rift zone, as evidenced by thermal springs at the coast and wells in the lower east-rift zone. Available data suggest that dike-impounded, heated ground water occurs at relatively high elevations in the upper east- and southwest-rift zones of Kilauea, and that permeability at depth in the rift zones (probably $\leq 10^{-15}$ m^2) is much lower than that of unaltered basalt flows closer to the surface ($\geq 10^{-10}$ m^2). Substantial variations in permeability and the presence of magmatic heat sources influence the structure of the fresh water–salt water interface, so the Ghyben-Herzberg model will often fail to predict its position. Numerical modeling studies have considered only subsets of the hydrothermal system, because no existing computer code solves the coupled fluid-flow, heat- and solute-transport problem over the temperature and salinity range encountered at Kilauea.

Key words: ground water, permeability, hydrothermal, volcanoes, Kilauea.

INTRODUCTION

In this paper we review and synthesize previous hydrogeologic studies of the Kilauea volcano area, and suggest possible directions for future research. Hydrogeologic investigations on the Island of Hawaii began in the late 19th century, when Kilauea volcano (Fig. 1) became the focus of a debate on the role of water in volcanic activity (Green, 1887; Brun, 1911; Day and Shepherd, 1913). Gaseous emanations from a tumulus in Halemaumau were sampled and analyzed in 1912–1913 (Day and Shepherd, 1913). The gases proved to be rich in CO_2, H_2O, and N_2, helping to disprove the view of Green (1887) and Brun (1911) that the magma was anhydrous.

Conventional hydrogeologic studies of Hawaii began in 1920, when the U.S. Geological Survey (U.S.G.S.) began studying the ground-water resources of the southern part of Hawaii, including Mauna Loa and Kilauea volcanoes (Fig. 1). O. E. Meinzer, head of the then Ground-Water Division of the U.S.G.S., summarized Hawaiian hydrology as follows:

> To a person who is familiar only with the mainland, the hydrologic conditions in the Hawaiian Islands seem to be freakish in every respect . . . The rainfall differs from anything found on the mainland in both intensity and areal variability . . . The islands consist of great volcanoes . . . Generally speaking, the lava rock is like a sieve . . . The hydrologic consequences of the widespread occurrence of this very permeable rock are obvious—rapid absorption and downward percolation of the rain water; small and flashy runoff; . . . great ground-water recharge and large supply of ground water; very low and flat water table with great depths to ground water in most places; copious springs in the low valleys, along the

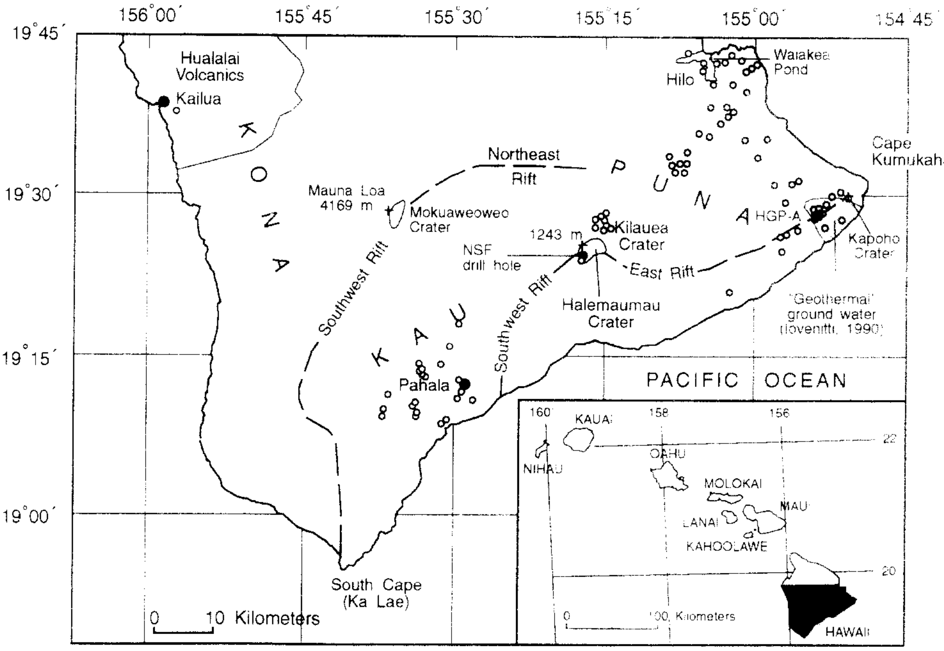

Fig. 1. Maps showing Kilauea volcano and adjacent parts of Mauna Loa volcano. Open circles indicate well locations.

coasts, and on the adjacent sea bottom; and salt water at no great depth below the water table.

O. E. Meinzer, in Stearns and Clark (1930)

Early U.S.G.S. and Hawaii Division of Hydrography studies emphasized the geologic setting and water-supply issues (Martin and Pierce, 1913; Stearns and Clark, 1930; Stearns and MacDonald, 1946), although Stearns and Clark (1930) also described historic and prehistoric phreatic eruptions (p. 143f) and related them to heated ground water (p. 155f). Stearns and Clark (1930) did not find any hot (≥37°C) springs in the Kau district (Fig. 1), but listed occurrences of steam vents and warm water (Stearns and Clark, 1930, p. 190). Stearns and MacDonald (1946) published the first extensive water-chemistry data for the Island of Hawaii (p. 287f) and presented comprehensive inventories of sea-level springs (p. 252f, 261f), wells (p. 285f), perched springs (p. 290f) and tunnels (p. 296f). They too mentioned the scarcity of hot springs, and attributed it to the high permeability of the volcanic rocks (p. 269). Davis and Yamanaga (1968) reported ground-water information from wells drilled between 1944 and 1967, and summarized the characteristics of streamflow records published by the U.S.G.S.

Since about 1970, hydrogeologic studies of the Kilauea area have focused on geochemistry. The most comprehensive presentation of geochemical data is that of Swain (1973), who reported water-chemistry data from 56 wells on Hawaii (p. 53f), including 21 wells in the South Hilo, Puna, and Kau districts (Fig. 1). Druecker and Fan (1976) mapped chloride and silica concentrations and temperatures for wells and springs in the Puna District. Data from deep geothermal wells have facilitated preliminary hydrologic and geochemical models of Kilauea's lower east-rift zone (Thomas, 1982, 1987; Iovenitti and D'Olier, 1985; Iovenitti, 1990).

DISTRIBUTION OF SPRINGS AND WELLS

Springs in the Puna and Kau districts are found near sea level and, less commonly, at higher elevations on the slopes of Mauna Loa. The sea-level springs frequently show tidal influences and their chemical composition is a mixture of fresh and salt water. Easily sampled springs are common and fairly evenly distributed along the south coast. Although historically documented springs are less common between Cape Kumukahi and Hilo (Fig. 1), thermal-infra-red imagery (Fischer *et al.*, 1966) indicates that there is also substantial ground-water discharge along the east coast. The relatively scarce springs at higher elevations on the slopes of Mauna Loa appear to discharge water from perched aquifers. Discharge rates from these springs fluctuate with rainfall. Water-supply wells in the Puna and Kau districts are concentrated in the Pahala area and between Hilo and Cape Kumukahi, generally within 15 km of the coastline (Fig. 1).

BASAL, PERCHED, AND DIKE-IMPOUNDED WATER

Hawaiian ground water has traditionally been classified as "basal", "perched", or "dike-impounded" (Fig. 2). This tripartite classification was well-established by 1920, when O. E. Meinzer visited Hawaii (Meinzer, in Stearns and Clark, 1930).

Basal ground water

In the late 19th century W. Badon Ghyben and A. Baurat Herzberg showed that the configuration of the "basal" fresh-water lens within permeable oceanic islands is a function of the relative density of salt and fresh water and of the altitude of the water table above sea level. If the water table stands 1 m above sea level (a.s.l.), then the salt water should be encountered 40 m below sea level (b.s.l.), under hydrostatic conditions and at standard temperature (25°C). The slope of the "basal" water table within the highly permeable Hawaiian basalts is generally very shallow; Peterson (1981, p. 11) cited an average value of 0.0002 for the Hawaiian Islands.

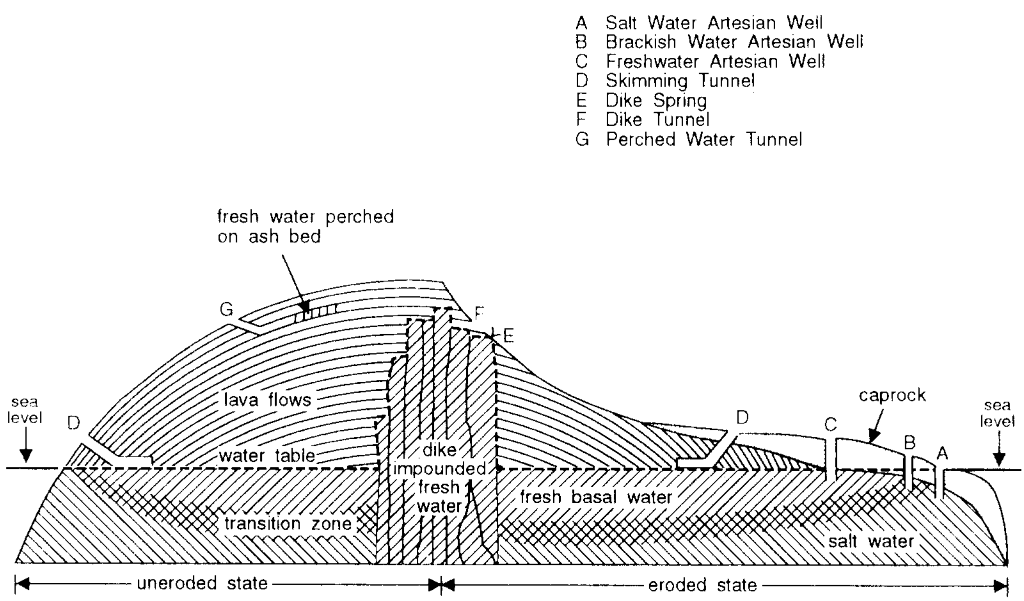

Fig. 2. Cross section of an idealized volcanic island showing occurrence and recovery of ground water (after Peterson, 1981, Fig. 4).

The average water-table gradient in the region between Hilo and Kilauea's east-rift zone is somewhat greater, about 0.001 (Stearns and MacDonald, 1946, p. 246). On the basis of these low gradients and Ghyben-Herzberg assumptions, Swain (1973, p. 32) concluded that the fresh-water lens beneath the Island of Hawaii cannot be thick. This conclusion has been disproved by later drill-hole data, which indicate that Ghyben-Herzberg is not a reliable model for the southern part of the island. For example, the 1966 m deep HGP-A well in Kilauea's lower east-rift zone (Fig. 1) produced low-salinity waters from depths as great as 1700 m b.s.l. (Kroopnick *et al.*, 1978, pp. 14–19), whereas the Ghyben-Herzberg model would predict only a 70 m thick fresh-water lens in the vicinity (Thomas, 1987, p. 1519). The land surface elevation is 194 m a.s.l. at HGP-A, and the water-table elevation in the vicinity is less than 2 m (Takasaki, 1993; Kauahikaua, this issue, Fig. 5).

Most basal ground waters that have been sampled are brackish sodium-chloride waters. Mixing with the underlying sea water is greatest near the shore, where the effects of tidal fluctuations are large. In areas where fresh-water recharge is small the entire ground-water lens may be brackish as much as 2 km from the shore (Davis and Yamanaga, 1973, pp. 29–30). Some wells tapping the upper part of the fresh-water lens produce sodium-bicarbonate waters (Swain, 1973, p. 32). Swain (1973) suggested that basal ground water from two wells with relatively high bicarbonate concentrations had been affected by volcanic activity. Chemical data from these wells are shown in Table 1. One is at Kapoho crater (lower east-rift zone), the other penetrates lavas of Hualalai volcano near Kailua (Fig. 1).

Perched ground water

"Perched" ground water is found within the unsaturated zone above the basal ground water (Fig. 2). The perching members are relatively impermeable ash layers or buried soils (e.g. O. E. Meinzer in Stearns and Clark, 1930, pp. 21–23). Perched water can be exploited by tunneling along the perching layer (Fig. 2) and has been extensively developed for agricultural and domestic use (Stearns and MacDonald, 1946, p. 254f, 263f, 296f). The Pahala Ash (Stearns and MacDonald, 1946, p. 68f) is an important perching layer on the southeast flank of Mauna Loa.

Dike-impounded ground water

Hawaiian volcanic dikes tend to be thin (averaging ~1 m) but can be laterally extensive. Scattered dikes can impound water in relatively large compartments of higher-permeability rock (Hunt *et al.*, 1988). The existence of "dike-impounded" water substantially above the basal level (Fig. 2) was first documented during construction of the Waiahole tunnel on Oahu in 1913–1915 (O. E. Meinzer, in Stearns and Clark, 1930, p. 23f). There, head differences of about 50 m were measured across 1–4 m thick dikes. Meinzer (in Stearns and Clark, 1930, p. 26) noted that the complexes of intersecting dikes that serve to impound ground water are found in rift zones along which magma has risen to feed eruptions.

Stearns and MacDonald (1946, pp. 224–225) suggested that concealed dikes in the interiors of Mauna Loa and Kilauea might impound ground water. The first direct evidence for impounded water on the Island of Hawaii was obtained in 1946, when an inclined shaft encountered the

Table 1. Chemical data from two water wells on Hawaii thought to be affected by volcanic activity

		(Concentrations in mg l^{-1})							
		pH	Ca	Mg	Na	K	HCO_3	SO_4	Cl
Kapoho crater	3/03/72	8.4	72	31	57	7.6	393	11	54
Hualalai volcanics	3/17/72	8.0	28	69	270	13	424	75	370

Table 2. Examples of dike-impounded waters

		(Concentrations in $mg\ l^{-1}$)							
		pH	Ca	Mg	Na	K	HCO_3	SO_4	Cl
Pahala	4/13/72	7.2	6.6	3.6	7.2	1.0	43	10	3.5
NSF drill hole on Kilauea's summit	12/18/74	8.2	49	64	769	142	150	1900	110
	6/06/75	7.6	72	87	730	111	792	1490	83

water table at 70 m elevation near Pahala (Fig. 1), where the basal water table would be expected to be only ~1.5–3 m a.s.l. Hussong and Cox (1967) used electrical resistivity soundings to demonstrate that the water level measured in the shaft correlated with a coherent low-resistivity layer (their Fig. 3) extending at least 1.2 km northeast of the well and 0.8 km south of the well. They (p. 5) suggested that dikes subparallel to the southwest-rift zone of Kilauea and (or) an unconformity act(s) to impound ground water in the Pahala area. In 1973, a National Science Foundation (NSF)-funded drill hole (Fig. 1) established that the water table beneath Kilauea summit is about 610 m a.s.l. (Keller *et al.*, 1979, p. 355). Subsequent resistivity line soundings suggested that water is also impounded to high elevations within Kilauea's east-rift zone (Jackson and Kauahikaua, 1987), and electrical soundings identified a continuous high-level low-resistivity layer that extends from Kilauea summit down the east- and southwest-rift zones (Jackson and Kauahikaua, 1990; Kauahikaua, this issue). In the Kilauea summit region, the top of the low-resistivity layer coincides with the water table. The layer abruptly deepens or disappears outside the rift zones; this might be due to a depth contrast (a deeper water table), a salinity contrast, or both.

Low-salinity sodium-bicarbonate waters sampled near Pahala (Fig. 1; Swain, 1973, well 1128-01) provide a representative example of perched or dike-impounded waters (Table 2). The relatively saline, bicarbonate- and sulfate-rich waters that underlie Kilauea's summit (Tilling and Jones, 1991a) and upper rift zones are atypical; their composition is presumably related to the ongoing volcanic activity. The composition of ground water below Kilauea summit was observed to change following the major eruptive and intrusive activity in the upper southwest-rift zone in December 1974: bicarbonate concentrations increased by a factor of five, and sulfate and chloride concentrations declined (Table 2; Tilling and Jones, 1991a, b).

PATTERNS AND RATES OF GROUND-WATER FLOW

Rates and patterns of ground-water flow in the Kilauea area are controlled by topography, the distribution of recharge, and the permeability structure. The presence of magmatic heat sources may also have a local effect on flow patterns.

There is no direct evidence for dike-impounded water within the rift zones of Mauna Loa (Fig. 1). However, zones of dike-impounded water probably exist, on the basis of analogy with Kilauea and other Hawaiian volcanoes (Swain, 1973). Thus the rift zones of Mauna Loa are probably the northern and western hydraulic boundaries of a regional ground-water flow system that includes Kilauea. The area bounded by the rift zones of Kilauea and the ocean (Fig. 1) may comprise a partly isolated subsystem.

The standard hydrogeologic assumption is that the water-table surface will be a subdued replica of the topography. In Hawaii, the presence of dike-impounded ground water causes deviations from this general expectation. Nevertheless, the general pattern of flow on the southern half of the island is from Mauna Loa summit towards the ocean. Flow in the shallow subsurface must accelerate below the middle-elevation belts of high rainfall ($\gtrsim 2.5\ m\ yr^{-1}$; Fig.

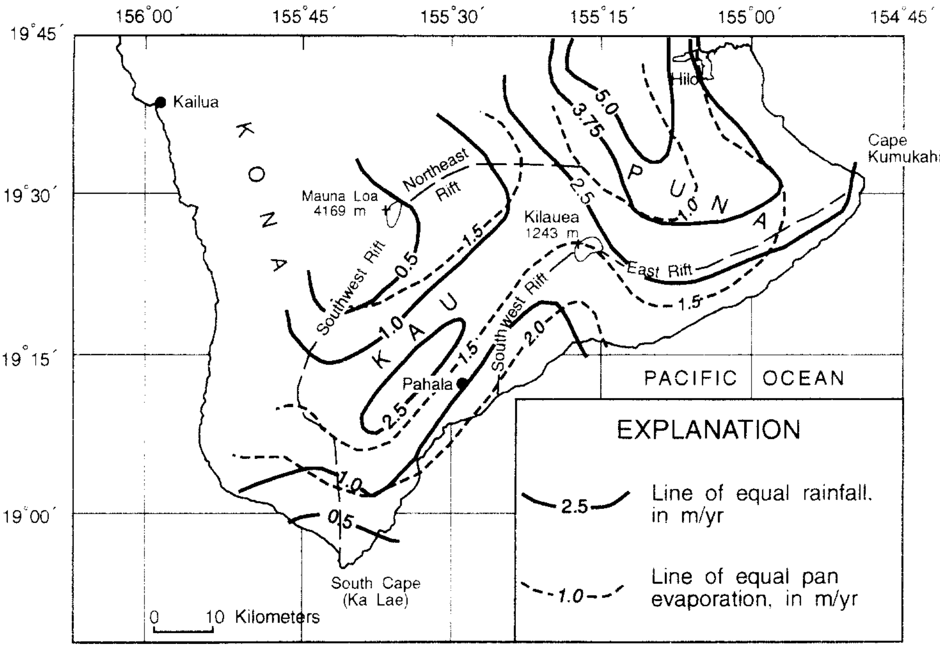

Fig. 3. Average annual rainfall and pan evaporation, Kau and Puna districts, Island of Hawaii. Rainfall contours from U.S.G.S. (1986), pan evaporation from Ekern and Chang (1985).

3). Because there is little or no surface runoff and potential evapotranspiration rates in these areas are ≤ 1.5 m yr^{-1} (Fig. 3), much of the rainfall is available to recharge the ground-water system.

Some ground-water flow beneath the flanks of Mauna Loa may be diverted by the rift zones of Kilauea. Dikes in Kilauea's lower east-rift zone appear to form a barrier to southward movement of ground water; basal ground water south of the rift zone is generally warmer and more saline than that to the north (Davis and Yamanaga, 1968, p. 30), and there is also a difference in basal water-table elevation across the east-rift zone (Druecker and Fan, 1976, p. 332). Takasaki (1993) stated that Kilauea's rift zones act to compartmentalize the ground-water flow system, so that recharge on the south flank of Mauna Loa above Kilauea does not reach the area bounded by the east- and southwest-rift zones of Kilauea and the ocean (Fig. 1). However, Imada (1984) successfully simulated water-table elevations in and near the lower east-rift zone by assigning relatively high permeabilities to rift-zone rocks (10^{-9} to 10^{-11} m^2).

Hydraulic relations within the rift zones are poorly understood. As noted above, evidence from the NSF drill hole and electrical soundings indicates a relatively high-level water table in the Kilauea summit area and adjacent portions of the upper east- and southwest-rift zones. Water-table elevations in the lower east-rift zone are close to sea level, but the HGP-A well (Fig. 1) produced dilute water from as much as 1.7 km b.s.l. If fresh or brackish ground waters exist at such depths in the lower east-rift zone, they may be relatively well-connected with dilute waters at higher elevations in the rift zone and poorly connected with saline waters outside the rift zone. Geothermal exploration wells in the lower east-rift zone have reportedly encountered large differences in pressure and salinity between nearby producing horizons, indicating the presence of major hydraulic barriers within the rift zone itself.

Large areal variations in the stable-isotope composition of meteoric recharge suggest that additional isotope data might lead to an improved understanding of ground-water flow patterns in Kilauea and adjacent areas. McMurtry *et al.* (1977) presented stable-isotope data for ground water from low-elevation wells. Additional data reported by Goff *et al.* (1991) showed deuterium and oxygen-18 values for cold meteoric waters ranging from 0 to −102‰ δD and 0 to −13.5‰ δ^{18}O, respectively, with areal variation controlled primarily by elevation differences. Scholl *et al.* (1992) documented a rate of decrease in precipitation δD values of 17‰ per 1000 m elevation: 92% of the variance in their data was explained by elevation differences.

Recharge and discharge

Total ground-water recharge on Kilauea and adjacent areas of Mauna Loa has been estimated at 150,000 l s^{-1} (Takasaki, 1993), or about 50–60% of total rainfall (Fig. 3). Water-balance studies of natural areas on the island of Oahu with similar average rainfall rates (2–2.5 m yr^{-1}) also suggest that about 50% of incident precipitation recharges the ground-water systems (Giambelluca, 1986, p. 40). There is substantial areal variation in recharge rates. Because surface runoff is negligible and actual evapotranspiration will not greatly exceed pan evaporation, recharge in the high-rainfall areas can be estimated as the difference between rainfall and pan evaporation. In the high-rainfall areas west of Hilo, recharge rates estimated on this basis exceed 4 m yr^{-1} (Fig. 3). In low-rainfall areas actual evapotranspiration is much less than pan evaporation, and recharge cannot be estimated by a simple differencing. Takasaki (1993) suggests recharge rates of 0.04–0.25 m yr^{-1} in areas where precipitation ranges from 0.5–1.5 m yr^{-1}. These values are consistent with those calculated by Giambelluca (1986) for the driest areas on Oahu.

Basal ground-water discharge is concentrated in strong flows only at a few places (Davis and Yamanaga, 1973, p. 31). The largest measured discharge from sea-level springs, in Hilo near Waiakea Pond (Fig. 1), is 5000 l s^{-1} (Hirashima, 1967), and Stearns and MacDonald's (1946) data indicate that the total yield of visible springs, tunnels, and wells on Hawaii amounts to only about 2.5% of the rainfall. They infer that huge unmeasured amounts of water issue from springs at or below sea level. This conclusion is consistent with later thermal-infra-red imagery of the Hawaiian coastline (Fischer *et al.*, 1966). The areas for which infra-red measurements were made show ubiquitous, plume-like thermal anomalies extending into the ocean (Fig. 4). These anomalies are presumably caused by ground-water discharge.

Although no hot springs have been reported to the south of the southwest-rift zone of Kilauea, the thermal imagery (Fig. 4) documents relatively warm spring-discharge temperatures along the shoreline below both the southwest- and east-rift zones. Outside the area bounded by the rift zones, spring temperatures are generally lower than the ambient ocean temperature. An apparent resistivity low (Flanagan and Long, 1987) and relatively high-temperature ground-water samples from two wells were used by Iovenitti (1990) to define an area of "geothermal" ground water south of the lower east-rift zone of Kilauea (Figs 1, 4). Iovenitti (1990) interpreted these data in terms of lateral migration of a thermal plume from an upwelling zone near HGP-A. The area of "geothermal" ground water includes the warm (~35°C) spring at Pohoiki reported by Druecker and Fan (1976, Table 3) and McMurtry *et al.* (1977, Table 1). However, the thermal imagery shows that, as of 1963, basal ground-water discharge temperatures in the area were generally lower than ocean temperatures, although it was bounded by two discrete areas of relatively warm discharge (Fig. 4).

PERMEABILITIES

Various numerical modeling results and sparse field observations permit some general inferences about the permeability structure of Kilauea. Permeability measurements are highly scale-dependent, and most values obtained from Hawaiian well tests and model results apply to

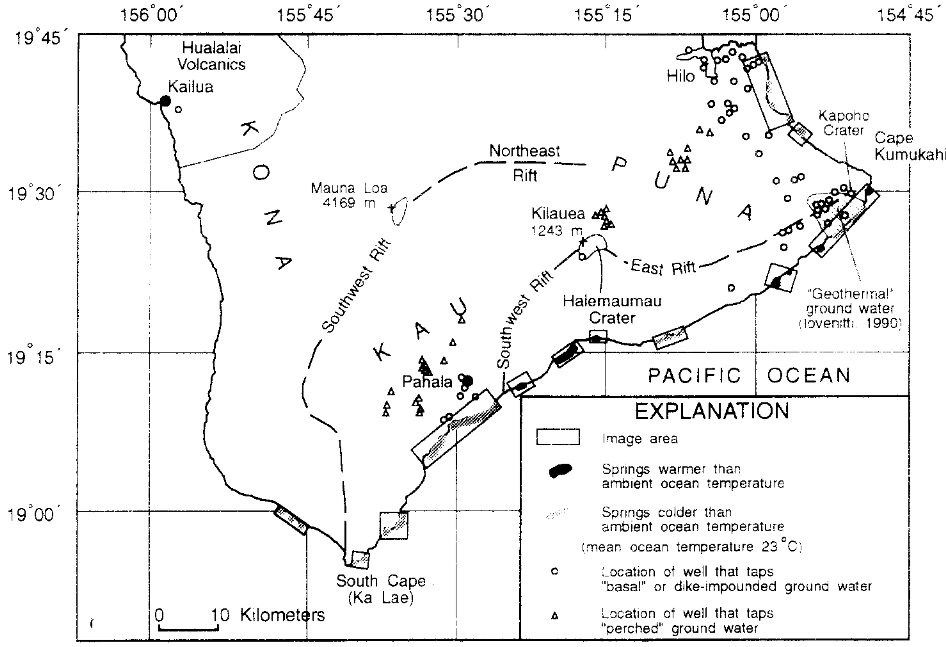

Fig. 4. Relative spring-discharge temperatures (Fischer *et al.*, 1966) and well locations (Takasaki, 1993) in Kau, Puna, and South Hilo districts, Island of Hawaii.

scales ranging from perhaps 100 m to tens of km. Over such scales, the permeability of Kilauea rocks probably ranges from $>10^{-10}$ m^2 to $<10^{-15}$ m^2, depending on the morphology of the lava flows, the frequency of dike intrusion, and the degree of hydrothermal alteration.

Imada (1984) simulated fluid flow and solute transport in a 20 km (NE–SW) by 12 km (NW–SE) area extending westward from Cape Kumukahi (Fig. 1). Her best match of observed water-table elevations was obtained by assigning permeability values of 5×10^{-9} m^2 to rocks outside of Kilauea's east-rift zone; 1×10^{-9} m^2 to rift-zone rocks within 10 km of Cape Kumukahi; and 1×10^{-11} to rift-zone rocks more than 10 km from the cape. The higher permeability of the eastern portion of the rift zone was attributed to decreasing dike density (Imada, 1984, p. 71). These values are similar to those assigned to unaltered, dike-free Hawaiian basalts in other modeling studies: Eyre (1985, p. 327) assigned a value of 2×10^{-10} m^2 to the Koolau aquifer of southern Oahu, and Voss and Souza (1987, p. 1861) assigned a value of 5×10^{-10} m^2 to the southern Oahu aquifer system. The average of 21 aquifer-test measurements of Hawaiian basalts reported by Williams and Soroos (1973) is a similar value, $1.8(\pm 1.4) \times 10^{-10}$ m^2. The measured permeability of the crust of Kilauea Iki lava lake (3×10^{-13} m^2) was substantially lower (Hardee, 1980, p. 217), but these *in situ* measurements sampled comparatively small volumes of rock.

Thus the horizontal permeability of dike-free, unaltered Hawaiian lava flows appears to be $\geq 10^{-10}$ m^2, and is comparable with that of karst limestone, well-sorted sand, or gravel (Freeze and Cherry, 1979, p. 29). The layering of lava flows introduces some anisotropy. Limited well-test data in Hawaiian lava suggest that vertical permeability is perhaps 10–100 times less than horizontal permeability (P. A. Hsieh, U.S. Geological Survey, pers. comm., 1991). Modeling studies, which estimate larger-scale permeabilities, have invoked somewhat larger degrees of anisotropy (e.g. Souza and Voss, 1987, p. 31).

Dike intrusion reduces horizontal permeability in the direction normal to the dike plane.

Where dikes are numerous and intersecting, overall horizontal permeability is substantially reduced (Takasaki and Mink, 1985, p. 7). Where ground-water temperatures are elevated, hydrothermal alteration also acts to reduce permeability. In the Cascade Range of the Pacific Northwest, substantial reduction in the permeability of volcanic rocks seems to be associated with temperatures ≥50°C (Blackwell and Baker, 1988, p. 59).

Murray (1974, pp. 56–61) estimated the permeability of the 500–1200 m depth interval in the NSF drill hole on Kilauea summit (Fig. 1) to be 1×10^{-14} m^2 on the basis of rates of mud loss during drilling. He inferred a value of 6×10^{-14} m^2 from numerical modeling experiments that simulated convection within a rectangular model and attempted to match the distinctive temperature profile (Fig. 5). Both values are significantly larger than the permeability of core samples obtained in that interval ($\sim 5 \times 10^{-17}$ m^2). Discrepancies of such magnitude ($\sim 10^3$) between core measurements and *in situ* or inferred values are commonly observed in fractured crystalline rocks (Brace, 1980, p. 247).

The reduced permeability in the lower part of the NSF drill hole appears to be related to alteration, rather than dike intrusion. The permeability of the unaltered rocks encountered in the upper part of the hole is relatively high. Reduced permeability (10^{-14}–10^{-13} m^2) below about 500 m depth is associated with partial filling of vesicles by secondary mineralization (Keller *et al.*, 1979, p. 353) and temperatures >60°C (Fig. 5). The mud-loss-based permeability value for the NSF drill hole (1×10^{-14} m^2) is an estimate of horizontal permeability, whereas the model-based value (6×10^{-14} m^2) assumes isotropic conditions. These estimates do not suggest

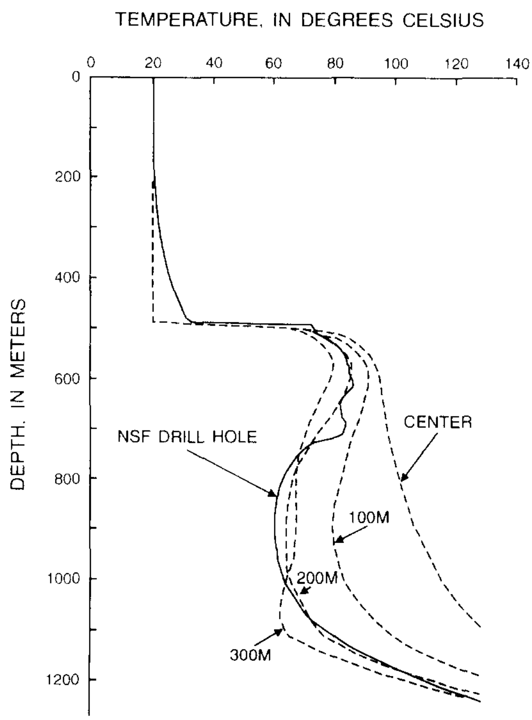

Fig. 5. Temperature profiles from Murray's (1974, Fig. 17) numerical simulations (dashed lines) compared with the observed profile from the NSF drill hole (solid line). Distances (100 m, 200 m, 300 m) are relative to the center of a model with a horizontal half-cell dimension of 700 m. The upper boundary is at 500 m depth and is held at a constant temperature of 20°C. The lower boundary is at 1900 m depth and maintained at 650°C. The 1200–1900 m depth interval is impermeable.

significant anisotropy; perhaps horizontal permeability has been preferentially reduced by loss of interflow and flow-top porosity.

The average permeability of the deep (500–1800 m b.s.l.) uncased interval in the HGP-A well (Fig. 1) is 2–4 × 10^{-16} m^2, on the basis of flow-test data (Chen *et al.*, 1978, 1979). These average values are obtained by dividing calculated transmissivities (permeability–thickness products) by the entire uncased thickness. Most of the production was attributed to relatively thin zones, which must have much higher permeability; the remainder of the interval must have lower permeability. Chlorite and montmorillinite, which occur interstitially and fill vesicles and fractures, are the dominant alteration minerals in this interval (Stone and Fan, 1978).

Well KS-8, near HGP-A in the lower east-rift zone, encountered fluid pressures of 134 bar at a depth of 1063 m (Thomas *et al.*, 1991). The land-surface elevation at KS-8 is 192 m and the water table in the vicinity is about 2 m above sea level, so this pressure exceeds local hydrostatic pressure by about 47 bar, or 470 m of cold-water head. A possible explanation for the high fluid pressures is that the well encounters a fracture zone containing high-pressure steam derived from a deeper, boiling horizon. Hedenquist and Henley (1985, Fig. 14) diagrammed the relevant pressure-depth relationships: if the pressure gradient in the fracture zone is nearly vaporstatic, then the 134-bar steam pressure would be close to that of the deeper, boiling fluid. Under hydrostatic boiling point-depth conditions, such pressures would be anticipated about 1800 m below the local water table, or about 2000 m below the land surface.

The high pressures at depth in KS-8 can alternatively be explained in terms of thermal pressurization, topographically driven flow, or both; either explanation would require low-permeability barriers between the 134-bar zone and the local water table. For thermal pressurization, confining permeabilities of $<10^{-16}$ m^2 would be required for a 50 bar pressure increase with an instantaneous temperature increase of 500°C and 1% porosity (Delaney, 1982, Table 12). Given 5% porosity, the same level of pressurization would be reached with a permeability $<10^{-15}$ m^2. For pressurization due to topographically driven flow, horizontal permeability between the upper east-rift zone or Kilauea caldera area (water table ≤610 m elevation) would have to be much greater than vertical permeability between the 134-bar zone and the local water table (2 m elevation). The required horizontal/vertical permeability ratio is likely larger than 10^2. Values of 10^2 or 10^3 are within the range of anisotropy values suggested by previous studies.

Temperature profiles from three other deep wells ("Scientific Observation Holes") in the lower east-rift zone of Kilauea suggest transitions from advection- to conduction-dominated conditions at depths varying from 1100 to 1400 m below sea level (Olson and Deymonaz, 1992; Kauahikaua, this issue, Fig. 8). Gradients at greater depths range from 200 to 320°C/km (Kauahikaua, this issue, Table 1), and temperatures range from 100 to 350°C. The apparent absence of large-scale convective circulation despite such high gradients implies bulk permeabilities $<10^{-14}$ m^2. Given a representative length of 500 m and temperature difference of 100°C, a permeability of 10^{-14} m^2 leads to a mean (250°C) Rayleigh number of about 100, well above the critical value for the onset of convection. The transition from advection- to conduction-dominated profiles at 1100–1400 m depth below sea level is not related to a particular isotherm, nor does it seem to be associated with the transition from subaerial to submarine basalts, which occurs ~560 m b.s.l. in the lower east-rift zone (SOH-1: Novak and Evans, 1991) and 1260–1400 m b.s.l. in the middle east-rift zone (SOH-4: Novak and Evans, 1991), where Kauahikaua (this issue, Table 1) defined a linear temperature gradient beginning ~1100 m b.s.l.

It is not clear whether pervasive intrusion or alteration is the primary cause of the relatively low bulk permeability at depth in the east-rift zone. Both are probably important, and related by the role of recently emplaced dikes as heat sources that help drive alteration. Thomas (1987, p. 1520) suggested that self-sealing by hydrothermal alteration may help isolate relatively low-salinity hydrothermal systems in Kilauea's lower east-rift zone from salt-water intrusion; anhydrite deposition by heated sea water was cited as a potentially important process. This mechanism has been documented in the Reykjanes high-temperature area of Iceland's axial rift zone. There, calcite and anhydrite deposition sustains pressure differences on the order of 10 bar across the boundaries of hydrothermal systems (Tomasson and Smarason, 1985, p. 197).

Ongoing tectonic activity may create localized zones of relatively high permeability at depth in the rift zones of Kilauea. Efforts to predict such fracture-related permeability in other areas have met with limited success. Most conventional geophysical techniques, with the possible exception of self-potential surveys, seem to be more effective at sensing porosity, which is often poorly correlated with permeability. Major fracture zones probably control advective heat and solute transport at depth and may allow exploitation of geothermal resources.

HYDROTHERMAL MODELING

Quantitative ground-water models have delineated some aspects of Kilauea's hydrothermal system (Cheng and Lau, 1974, 1978; Murray, 1974; Lau and Cheng, 1977; Cheng and Teckchandani, 1977). For instance, Murray (1974, p. 49f) and Cheng and Lau (1978, p. 10) showed that the distinctive temperature profile from the NSF drill hole at Kilauea summit (Fig. 5) can be explained in terms of free convection in a confined aquifer. Murray's best match to the temperature profile involved an impermeable upper boundary near the water table (~500 m depth) and an impermeable zone of conductive heating below 1200 m depth.

Cheng and Lau (1974) simulated free convection in an unconfined geothermal reservoir and Lau and Cheng (1977) considered the thermal effects of dike intrusion into an unconfined reservoir. Both studies illustrate how the distribution of heat sources might affect the thermal field in the reservoir. Lau and Cheng (1977, p. 1209) suggested that the water table rises in response to dike emplacement. This result was due to thermal expansion of the fluid and should be considered semi-quantitative, because some factors that could influence water-table response were neglected: fluid viscosity was assumed constant, and the mechanical effects of dike emplacement were not treated.

Future directions

Like most modeling studies of fluid flow near intrusions (Furlong *et al.*, 1991, p. 473), the modeling studies of Kilauea have assumed that the circulating fluid is pure water. The Kilauea studies also involved idealized, two-dimensional geometries and invoked certain simplifying assumptions. Most important, fluid properties were assumed to be constant, except for fluid density in the buoyancy force term (the Boussinesq approximation), and the temperature of the fluid was assumed to be below the boiling-point temperature for the ambient pressure.

Some computer simulation codes avoid these approximations, at least for the case of subcritical ($\leq$350°C) pure-water systems (e.g. Faust and Mercer, 1979a, b), and some models can simulate fluid flow with coupled heat and solute transport at relatively low ($\leq$100°C) temperatures (e.g. Kipp, 1987). TOUGH2 (Pruess, 1991) is currently the state-of-the-art hydrothermal simulation model, combining solutions to fluid-flow, heat- and solute-transport equations. However, even TOUGH2 is limited to subcritical temperatures, and the solute-transport algorithm is not yet documented. No existing code solves the coupled problem over the entire temperature (0–1200°C) and salinity range (perhaps 0–20% NaCl: Fournier, 1987) of a Kilauea-like system (Fig. 6). Further, field observations seem to require that comprehensive models of the hydrothermal system account for the coupling between deformation and fluid flow and transport. In tectonically active environments, deformation affects permeability (Brenan, 1991) and is also a driving force for fluid flow (Elsworth and Voight, 1992). Geoscientists cannot

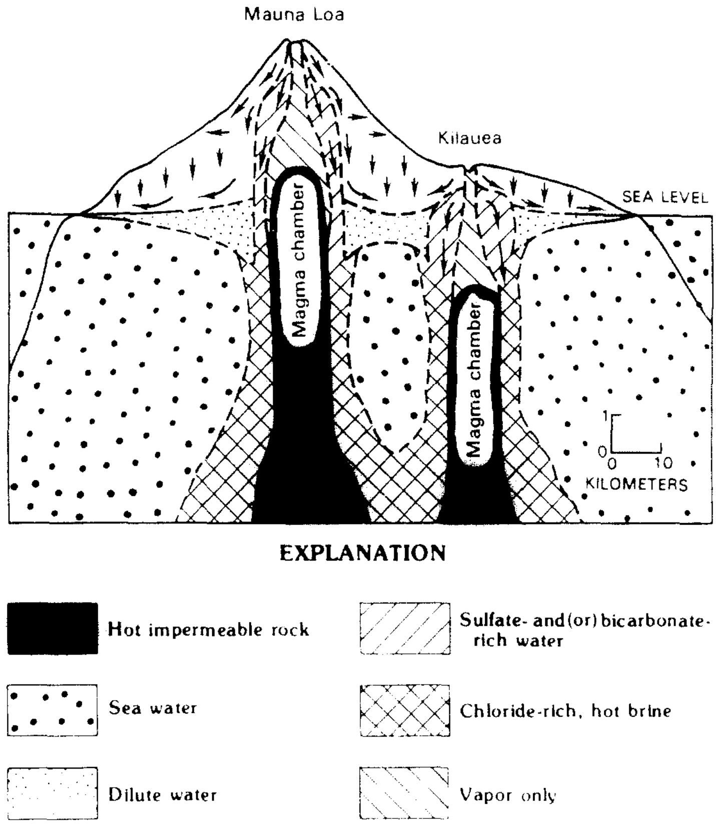

Fig. 6. Conceptual model of shallow magma reservoirs and associated hydrothermal systems beneath Mauna Loa and Kilauea volcanoes (from Fournier, 1987, Fig. 55.15). Arrows indicate inferred direction of water movement.

yet simulate the entire set of relevant processes. Computational capability is inadequate, and quantitative descriptions of the physical parameters controlling each process is incomplete. However, results of both generic (Forster and Smith, 1988, 1989) and site-specific studies (Norton and Taylor, 1979; Ingebritsen and Sorey, 1985; Willet and Chapman, 1987; Ingebritsen *et al.*, 1991) imply that simulation of a subset of the coupled processes can provide useful insight into transport phenomena.

The work of Forster and Smith (1988, 1989) is perhaps the most relevant to the understanding of Kilauea, because it deals with flow and heat transport in an unconfined, mountainous flow system. These authors illustrated some of the factors controlling water-table configurations and temperature fields; for example, Fig. 7 shows that a two-fold increase in basal heat flux would cause a substantial decrease in water-table elevation, due largely to decreased fluid viscosity at higher temperature. Forster and Smith (1988, 1989) dealt with relatively low-temperature systems, and general insight into higher-temperature ground-water flow near intrusions has been provided by other authors, for example Norton and Taylor (1979) and Lister (1983).

Questions that might be addressed in the course of future quantitative modeling studies of the hydrogeology of Kilauea include:

What is the position and nature of the salt water–fresh water interface beneath Hawaii? How is it affected by heat sources and variations in permeability?

What pressure, temperature, and fluid-saturation conditions are found between magma and

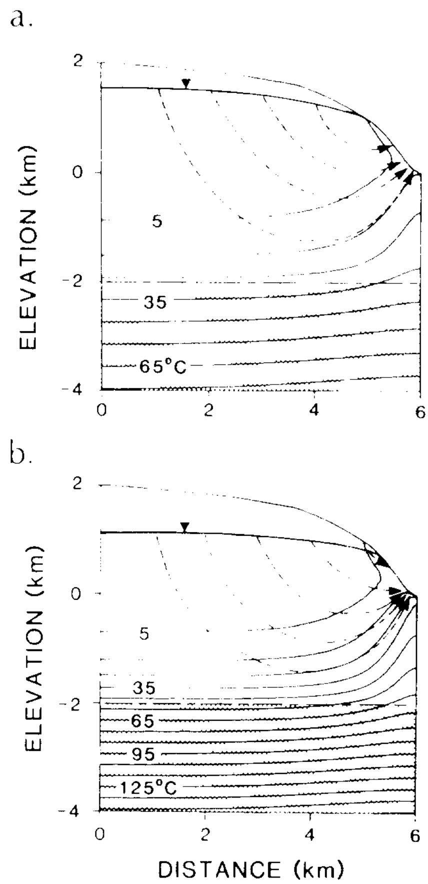

Fig. 7. Influence of basal heat flux on water-table configuration, fluid-flow pattern, and temperature field: (a) basal heat flux = 60 mW m^{-2}; and (b) basal heat flux = 120 mW m^{-2} (from Forster and Smith, 1988, Fig. 10). In each case the permeability of the upper (unshaded) unit is 1×10^{-15} m^2 and the infiltration rate is 2×10^{-9} m s^{-1}. Heavy line indicates the water table and dashed lines are flow lines. The lateral boundaries are closed and insulated; the lower boundary is closed, with a prescribed heat flux; and the water table is treated as a free surface. Fluid infiltration between the land surface and water table is one-dimensional (vertical) and the temperature of the infiltrating fluid is controlled by land surface elevation.

the ground surface? How are these conditions related to heat and mass transfer between magma and the surface?

What are the causes of decreased permeability at depth in the rift zones? What is their relative importance?

How might spatial/temporal variations in permeability due to geochemical or mechanical processes influence the evolution of the hydrothermal system?

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